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Estimating volumetric surface moisture content for cropped soils using a soil wetness index based on surface temperature and NDVI

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ABSTRACT

Surface soil wetness determines moisture availability that controls the response and feedback mechanisms between land surface and atmospheric processes. A study was carried out to estimate volumetric surface soil moisture content (θ_v) in cropped areas at field ($<10^2$ m) to landscape ($\leq 10^3$ m) scales. Triangular scatters from land surface temperature (LST) and normalized difference vegetation index (NDVI) space were utilized to obtain a soil wetness index (SWI), from which θ_v was derived, with the combination of dry and wet edges using data from ASTER (Advanced Space borne Thermal Emission and Reflection Radiometer) for field scale and MODIS (MODerate resolution Imaging Spectroradiometer) AQUA for landscape scale studies. The root mean square error (RMSE) of field scale θ_v estimates was higher (0.039 $m^3\,m^{-3}$) than that of the landscape scale (0.033 $m^3\,m^{-3}$). The narrow swath ($\sim$60 km) of finer resolution sensors (e.g. ASTER) often fails to capture the surface heterogeneity required in the triangle method for deriving SWI and could be one of the main reasons leading to relatively high error in θ_v estimates. At both the scales, the lowest error of θ_v estimates was found to be restricted within the NDVI range of 0.35–0.65. A geostatistical technique was applied to assess the influence of sub-pixel heterogeneity as an additional source of error for cross-scale comparison of θ_v estimates obtained from LST–NDVI scatters. The overall errors of θ_v estimates from LST–NDVI space were comparable with other globally available test results. The comparison of landscape scale θ_v from MODIS AQUA with large-area global estimates from a passive microwave sensor (e.g. AMSR-E) with longer microwave frequency (e.g. C-band) yielded 75% correlation and 0.027 $m^3\,m^{-3}$ root mean square deviation (RMSD) for fractional vegetation cover less than 0.5. The study recommends the synergistic use of shorter microwave frequency (e.g. L-band) and optical–thermal infrared bands as the best way of satellite based passive soil moisture sensing for vegetated surfaces.

1. Introduction

Near-surface soil moisture (0–0.05 m) is an important hydrological variable influencing the interactions between the land surface and atmospheric processes (Brubaker and Entekhabi, 1996). The partitioning of energy between sensible and latent heat fluxes is strongly controlled by surface wetness (Monteith, 1981; Shuttleworth, 1991; Entekhabi, 1995; Cahill et al., 1999; Dirmeyer et al., 2000). Through this control, soil moisture influences local weather variables, including boundary layer height and cloud coverage (Betts and Ball, 1998). It is important to understand these control and feedbacks when using global circulation models (GCMs) (Beljaars et al., 1996), regional climate models (Wetzel and Chang, 1988; Wilson et al., 1987), crop growth simulation models (de Wit and van Diepen, 2007; Guerif and Duke, 2000) and rainfall-runoff transformation models (Eltahir, 1998), for example. Despite the multidimensional importance of surface soil moisture, reliable regional and regular determination of this variable is extremely difficult through conventional point measurements, which are complex, expensive and available at a limited number of stations only.

A number of satellite-based techniques are in use for spatiotemporal soil moisture sensing. These utilize the electromagnetic spectrum in the optical–thermal and microwave bands. The latter has all-weather surface sensing capability in contrast to the former, which can sense only in clear skies. Passive microwave sensors (e.g. SSMI, AMSR-E) have 1–1.5 passes per day but have a coarse spatial resolution ($\sim$25 km). Therefore, the brightness temperature data from this type of sensors could be converted only to estimate large-area soil moisture as shown in a number of long term soil moisture mapping experiments such as SMEX02 (Cosh et al., 2004; Jacobs et al., 2004), SMEX03 (Jackson et al.,

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2005), SMEX04 (Vivoni et al., 2008), SMOSREX (de Rosnay et al., 2006), SMACEX (Kustas et al., 2005) and soil moisture mapping missions like SMOS (Kerr et al., 2001), AMSR-E (Njoku and Li, 1999; Njoku et al., 2003), *IRS-P4* MSMR (Thapliyal et al., 2005), SSMI (Basist et al., 2001; Singh et al., 2005). However, active microwave sensors (SAR: Synthetic Aperture Radar) have higher spatial resolution (10–30 m), yet have low repeat intervals (once in 16–25 days). It is also known that SAR (e.g. ERS, ENVISAT) backscatter data in higher frequency (e.g. C-band) provide less accurate surface soil moisture estimates for vegetated or cropped fields than for bare ones.

Short-term variation of surface soil moisture in vegetated landscapes with greater spatial details can be frequently monitored only through optical–thermal infrared band data from moderate resolution (250 m to 1 km) sensors (e.g. MODIS TERRA and AQUA) having two overpasses per day. Sub-pixel soil moisture estimates in vegetated surface within a landscape could be obtained using data from similar bands with finer resolution (<100 m) satellite sensors such as Landsat or ASTER. Normalized difference vegetation index (NDVI), derived from red and near-infrared (NIR) bands, alone is not sensitive to short time soil moisture variations (Fensholt and Sandholt, 2003). However, the NIR-shortwave infrared (SWIR) based indices such as SIWSI (Fensholt and Sandholt, 2003), SASI (Khanna et al., 2007) have been found to capture soil wetness at 1 km resolution. The concept of using data from the thermal infrared (TIR) band to monitor canopy water stress was originally proposed by Jackson et al. (1977) who developed the crop water stress index (CWSI). A number of studies (Sandholt et al., 2002; Carlson, 2007; Nemani et al., 1993) have suggested that the combined information from land surface temperature (LST) and NDVI can provide better information on vegetation stress and moisture conditions at the surface. As a given area dries, it is expected that the negative relationship between NDVI and LST will be altered due to increased LST for the low NDVI areas. This leads to a triangular space that has been explored rigorously to extract surface soil moisture status (Carlson et al., 1995; Gillies and Carlson, 1995; Nemani et al., 1993; Smith and Choudhury, 1991; Goetz, 1997; Sandholt et al., 2002). An overview of the LST–NDVI 'triangle method' for estimating soil surface wetness can be found in Carlson (2007).

Surface soil moisture content can also be potentially derived from estimates of surface thermal inertia. The feasibility of using apparent thermal inertia (Rosema and Fiselier, 1990; Stisen et al., 2008) from noon–night LST difference or morning rise in LST to estimate surface soil moisture at coarser scales (2–8 km) was demonstrated (Wetzel and Woodward, 1987; Hayden et al., 1996; Verstraeten et al., 2006) using optical and thermal data from geostationary sensors. This technique has limitations due to variable cloud cover conditions between day and night or two different hours within daytime. The most convenient approach would be to extract surface soil wetness using optical-TIR band data from the same overpass. The LST–NDVI space was, therefore, utilized in the present study to derive surface soil wetness (i) to estimate volumetric soil moisture content followed by validation with *in situ* measurements in cropped soils at both field (<10^2 m) and landscape ($\leq 10^3$ m) scales; (ii) to compare soil moisture estimates at these two different scales and (iii) finally to compare landscape scale estimates with global large-scale soil moisture from passive microwave sensor.

2. Surface temperature–vegetation index space

2.1. Theory and hypothesis

Land surface temperature largely depends on soil moisture and fractional vegetation cover. However, no universal and direct relation between LST and soil moisture is evident. Several workers (Stisen et al., 2008; Wang et al., 2006) demonstrated the physical relationship between vegetation index (e.g. NDVI) and LST in the form of a scatterplot, which could be a triangular variant of trapezoidal space. This implies that the sensitivity of LST to soil moisture differs for the canopy and for the soil surface beneath the plants. This sensitivity to soil moisture content tends to be much greater for bare soil than for canopies. This association between LST and NDVI, therefore, allows estimation of soil moisture content.

The following mechanisms are suggested as those determining the location of a pixel in the LST–NDVI space. These are based on the earlier findings reported in the literature (Sandholt et al., 2002; Carlson, 2007). The fractional vegetation cover can be related to spectral vegetation indices such as NDVI. This influences the amount of bare soil and vegetation visible to a sensor and differences in radiative temperature between the soil and canopy will affect the spatially integrated LST. Evapotranspiration largely controls the surface temperature as it is a term in the surface energy balance, which stipulates that

$$R_n = \mathrm{LE} + H + G \tag{1}$$

where R_n is the net radiation, LE is the evapotranspiration, H is the sensible heat flux and G is the soil heat flux (all fluxes in W m^{-2}). The surface temperature plays a key role in all of these fluxes.

The lower LE, the more energy available for sensible heating of the surface, H. Stomatal resistance to transpiration is a key factor, which is partly controlled by soil moisture availability. Thermal inertia (a thermal soil property which combines heat capacity and thermal conductivity, see e.g. Murray and Verhoef, 2007) influences LST in the case of partly vegetated surfaces, as they affect the amount of heat transferred into the soil, G. These thermal properties are a function of soil type and change with surface soil moisture (Verhoef et al., 1996). The available energy incident at the surface ($R_n - G$) also affects LST. The radiative control of LST implies that areas with a lower net shortwave radiation balance (e.g. due to a high albedo) will have lower temperatures, all else equal. Albedo is controlled by soil type, surface soil moisture, and vegetation cover. Incident radiation also affects the stomatal resistance to transpiration, which factors into the partitioning of net radiation into sensible and latent heat. The ability to transfer heat away from the surface into the atmosphere is an important component in the control of LST. This, in part, explains how vegetated surface with higher roughness length have lower LST (all else equal) compared to bare soil, which influences the shape of the LST/NDVI space. Similarly, homogeneous surfaces with unlimited water supply (for instance, irrigated surfaces) may have higher LST than expected, if turbulent exchange is reduced by poor mixing (Nemani and Running, 1997).

The conceptual LST–NDVI triangular space, with LST (y-axis) plotted as a function of NDVI (x-axis) is shown in Fig. 1. The hypotenuse of the triangle represents a warm/dry edge consisting of a group of points indicative of zero surface soil wetness for different NDVI classes in contrast to the cold/wet edge represented by the base of the triangle having maximum surface soil wetness conditions at different NDVI classes. The left edge represents bare soil ranging from dry to wet (top-down). For bare soils, at constant irradiance, LST is primarily determined by soil moisture content, via evaporative control and thermal properties of the surface. As the green vegetation increases along the x-axis (with NDVI), the maximum LST (=Ts_{max}) decreases. For dry conditions, the negative relation is defined by the warm/dry edge, which is the upper limit of LST for a given surface type and climate forcing. The 'wet' edge consists of a group of points